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AFGL-TR-76-0127
ENVIRONMENTAL RESEARCH PAPERS, NO. 566

# Rise of Volcanic Eruption Clouds: Relationship Between Cloud Height and Eruption Intensity

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22 June 1976

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TERRESTRIAL SCIENCES DIVISION PROJECT 8607
AIR FORCE GEOPHYSICS LABORATORY
HANSCOM AFB, MASSACHUSETTS 01731

AIR FORCE SYSTEMS COMMAND, USAF

This technical report has been reviewed and
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FOR THE COMMANDER:

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Unclassified

SECURITY CLASSIFICATION OF THIS PAGE (When Data Entered)

| REPORT DOCUMENTATION PAGE | | READ INSTRUCTIONS BEFORE COMPLETING FORM |
|---|---|---|
| 1. REPORT NUMBER AFGL-TR-76-0127, AFGL-ERP-566 | 2. GOVT ACCESSION NO. | 3. RECIPIENT'S CATALOG NUMBER |
| 4. TITLE (and Subtitle) RISE OF VOLCANIC ERUPTION CLOUDS: RELATIONSHIP BETWEEN CLOUD HEIGHT AND ERUPTION INTENSITY. | | 5. TYPE OF REPORT & PERIOD COVERED Scientific. Interim. |
| | | 6. PERFORMING ORG. REPORT NUMBER ERP No. 566 |
| 7. AUTHOR(s) Mark Settle 1st Lt, USAF | | 8. CONTRACT OR GRANT NUMBER(s) |
| 9. PERFORMING ORGANIZATION NAME AND ADDRESS Air Force Geophysics Laboratory (LWW) Hanscom AFB Massachusetts 01731 | | 10. PROGRAM ELEMENT, PROJECT, TASK AREA & WORK UNIT NUMBERS 86070501 61102F 05 |
| 11. CONTROLLING OFFICE NAME AND ADDRESS Air Force Geophysics Laboratory (LWW) Hanscom AFB Massachusetts 01731 | | 12. REPORT DATE 22 June 1976 |
| | | 13. NUMBER OF PAGES 37 |
| 14. MONITORING AGENCY NAME & ADDRESS(if different from Controlling Office) Environmental research papers | | 15. SECURITY CLASS. (of this report) Unclassified |
| | | 15a. DECLASSIFICATION DOWNGRADING SCHEDULE |

14 6 10 16 17 11 9 12

16. DISTRIBUTION STATEMENT (of this Report)

Approved for public release; distribution unlimited.

35p.

17. DISTRIBUTION STATEMENT (of the abstract entered in Block 20, if different from Report)

18. SUPPLEMENTARY NOTES

19. KEY WORDS (Continue on reverse side if necessary and identify by block number)

Volcanic eruptions
Cloud rise
Stratosphere
Atmospheric dust
Thermal plumes
Infrared sources

20. ABSTRACT (Continue on reverse side if necessary and identify by block number)

The rise of eruption clouds is produced by the upward momentum and thermal buoyancy of volcanic dust and gas, ~~These~~ processes, which play important roles in other phenomena. The expansion of a turbulent jet in free flow, (that is, unconfined by lateral boundaries) is controlled by the rate at which the forward momentum of the jet is dissipated. ~~The~~ thermal buoyancy of industrial waste gases provides a mechanism for moving such waste ~~materials~~ upward through the atmosphere and ensuring their dispersal over a wide area. The rise of

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DD FORM 1 JAN 73 1473 EDITION OF 1 NOV 65 IS OBSOLETE

Unclassified

SECURITY CLASSIFICATION OF THIS PAGE (When Data Entered)

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20. Abstract (Continued) ?fr p 1473A)

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volcanic eruption clouds can be modelled after these ~~two~~ analogous phenomena. In this report average ejection velocities ~~($w_o$)~~ at a volcanic vent ranging from 20 m/sec to 200 m/sec are assumed to represent a wide range of eruption intensity, from Strombolian to Vulcanian types, ~~of eruption.~~ For eruption velocities varying from 20 m/sec to 200 m/sec, cloud heights estimated by the turbulent jet model range from 1500 m to 6500 m (mid-latitude eruption) while cloud heights estimated by the industrial plume models range from 900 m to 10,000 m. These estimates are considered to be roughly comparable in view of the assumptions and extrapolations involved in applying these models to explosive eruption conditions and agree quite well with reported heights of eruption clouds. The fact that comparable estimates of cloud height are produced by ~~the two~~ very different models suggests that both momentum and thermal buoyancy play an important role throughout the main portion of an eruption cloud's trajectory. For these eruption conditions (20 m/sec $\leq w_o \leq$ 200 m/sec), neither momentum nor thermal buoyancy appears to dominate the process of cloud rise to altitudes of $\sim$ 10 km above an actively erupting volcanic vent. An order of magnitude variation in eruption velocity from $w_o$ = 20 m/sec to $w_o$ = 200 m/sec results in a factor of 3 to 4 increase in average cloud height predicted by the turbulent volcanic jet model and a factor of 2.5 increase in median cloud height predicted by a select group of industrial plume models. However, both models also demonstrate that changes in crosswind velocity by factors of 2 to 5 can result in variations in cloud height of similar magnitude. Therefore, reported heights of eruption clouds without reference to local crosswind conditions at the time of an eruption cannot be directly compared to gauge the relative explosiveness of different volcanic eruptions.

1473B

## Preface

The author is grateful to Tom Webb, Chuck Wood, and John Cronin for critically reviewing earlier versions of this report, and to Elaine Robson for her effort and patience in preparing the manuscript.

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# Rise of Volcanic Eruption Clouds: Relationship Between Cloud Height and Eruption Intensity

## I. INTRODUCTION

Explosive volcanic eruptions inject large quantities of ash and gas into the earth's atmosphere. The length of time these different volcanic products reside in the atmosphere can vary from several hours to several years. As a result, an individual eruption can produce meteorological effects that range in time from several days to several years and can range in space from a localized region to the entire planet.

Regional meteorology can be significantly altered by a major explosive eruption. Airborne ash and volcanic gases can effectively insulate the earth's surface, modifying diurnal temperature variations and producing a short-term warming of the region. Rainwater from clouds contaminated with volcanic gases can be highly acidic and may pollute local ground water.

The long term atmospheric effects of an eruption are produced by particulate dust and gases that have much longer atmospheric residence times. Major explosive eruptions can have a significant impact on the chemical budget and radiation budget of different portions of the atmosphere. Volcanic eruptions appear to be the dominant source of atmospheric chlorine[1] which plays an important role in

(Received for publication 22 June 1976)

1. Ryan, J.A., and Mukherjee, N.R. (1975) Revs. Geophys. and Space Sci. 13:650-688.

ozone chemical reactions in the stratosphere.[2] Sulphur dioxide gas produced by volcanic eruptions is a relatively minor source of atmospheric sulphur[3] but can significantly increase the density of the stratospheric aerosol layer by gas phase oxidation to sulphate particles.[4, 5, 6] The injection of large quantities of silicate dust particles and sulphur gases into the atmosphere can produce a cooling of the earth's surface by increasing global albedo or an increased greenhouse warming of the surface due to the opacity of these volcanic products to infrared radiation emitted from the earth's surface. Theoretical calculations of Pollack et al[7] indicate that as larger dust particles settle out of the atmosphere, global cooling becomes the predominant effect.

The atmospheric impact of a particular eruption is largely determined by the altitudes at which volcanic dust and gas enter and are mixed into the atmosphere. Precipitation in the troposphere effectively washes these materials out of the lower atmosphere. Tropospheric weather systems also mix large air masses over relatively short periods of time, rapidly reducing the concentration of volcanic products. The lower boundary of the troposphere is the earth's surface which provides a variety of geological, biological and anthropogenic sinks for airborne volcanic products. Long term atmospheric effects from individual eruptions are therefore limited to eruptions that succeed in penetrating the upper levels of the troposphere and introduce volcanic dust and gas into the stratosphere.[8, 9] The average height of the tropopause varies latitudinally from approximately 9 km at the poles to approximately 16 km at the equator.

The rise of an eruption cloud is controlled by the upward momentum of ash and gas at the mouth of a volcanic vent and by the thermal buoyancy of the volcanic gases. The initial rise of dust and gas in an eruption cloud is largely determined by the exit velocity of the material. At higher altitudes the initial momentum of the volcanic dust and gas has been substantially dissipated and the subsequent rise of the eruption cloud is predominantly determined by the relative buoyancy of the hot volcanic gases. This transition is sometimes reflected in the morphology of the

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2. Rowland, F.S., and Molina, M.J. (1975) Revs. Geophys. and Space Sci. **13**:1-35.
3. Kellogg, W.W., Cadle, R.D., Allen, E.R., Lazarus, A.L., and Martell, E.K. (1972) Science **175**:587-596.
4. Harker, A.B. (1975) J. Geophys. Res. **24**:3399-3401.
5. Lazrus, A.L., and Gandrud, B.W. (1974) J. Geophys. Res. **79**:3424-3431.
6. Dyer, A.J. and Hicks, B.B. (1968) Quart. J. Roy. Meteorol. Soc. **94**:545-554.
7. Pollack, J.B., Toon, O.B., Sagan, C., Summers, A., Baldwin, B., and Van Camp, W. (1976) J. Geophys. Res. **81**:1071-1083.
8. Lamb, H.H. (1970) Phil. Trans. Roy. Soc. London **266**:425.
9. Cronin, J.F. (1971) Science **172**:847-849.

eruption cloud. The lower portion of the cloud contains a larger concentration of solid ejecta and can appear much darker than the upper portion of the cloud. In such cases the lower portion of the cloud is sometimes referred to as the ash cloud, whereas the upper, lighter-colored portion is sometimes termed the vapor cloud. In other instances eruption clouds have a uniform grey appearance. Reported heights of eruption clouds generally refer to the maximum height of the condensed vapor cloud observed above an actively erupting volcanic vent. Significant amounts of particulate dust and gas may actually rise beyond the top of the observable cloud.

The maximum height of an eruption cloud is related to the intensity of the explosive eruption. The most intense explosive eruptions are characterized by ejection velocities on the order of hundreds of meters per second and large mass flux rates. In the past the classification of different styles of explosive eruption has been qualitatively based upon a variety of parameters, including the viscosity and chemical composition of the erupted magma, and the violence of a particular eruption measured in terms of loss of life or the extent of property destruction.[10] Most classification schemes include a general description of the size and structure of the eruption cloud associated with a particular type of explosive eruption. Such descriptions suggest that the size of an eruption cloud is approximately correlated with eruption intensity, with small clouds, rising to heights of several hundred meters, associated with weakly-explosive Strombolian-style eruptions, and larger clouds, rising to heights of several kilometers, associated with violently-explosive Vulcanian-style eruptions. Thus the height of an eruption cloud can be considered to be an approximate index of eruption intensity. Reports of the heights of eruption clouds observed in remote areas, where ground-based observations of active eruptions are hazardous or impossible, have been used to qualitatively gauge the relative intensity of such eruptions.

## 2. PURPOSE OF THIS STUDY

The physical processes which are responsible for the rise of eruption clouds — the upward momentum and thermal buoyancy of the erupted material — play important roles in other phenomena. The expansion of a turbulent jet in free flow (that is, unconfined by lateral boundaries) is controlled by the rate at which the forward momentum of the jet is dissipated. The structure of turbulent jets is a classical

10. MacDonald, G.A. (1972) Volcanoes, Prentice-Hall, Englewood Cliffs, N.J., 510 pp.

problem in fluid mechanics.[11] The thermal buoyancy of industrial waste gases provides a mechanism for moving such waste materials upwards through the atmosphere and ensuring their dispersal over a wide area. The release and dispersion of such industrial effluents in the atmosphere have been described by a wide variety of theoretical and empirical studies (see, for example, summary by Briggs[12]). The rise of volcanic eruption clouds can be modelled after these two analogous phenomena. Such models are developed in this study to investigate the relationship between eruption cloud height and exit conditions at a volcanic vent. The purpose of this study is twofold: (1) to determine if the rise of eruption clouds is predominantly controlled by the initial momentum or thermal buoyancy of volcanic products; and (2) to determine if the heights of eruption clouds are an accurate reflection of relative eruption intensity.

## 3. ERUPTION CLOUD RISE ESTIMATES

### 3.1 Turbulent Jet Flow in the Atmosphere

Descriptions of the structure of eruption clouds have been made principally by ground-based observers who have reported the shape and size of such clouds. Highly variable winds, large quantities of particulate ash, and occasional electrical storms associated with eruption clouds make aerial observations difficult (see, for example, Thorarinsson and Vonnegut[13]). As a result, very little is known about the internal structure of eruption clouds or about variations in local meteorological conditions (for example, temperature gradients, humidity, or wind structure) in the vicinity of eruption clouds.

An approximate model of the internal structure of eruption clouds may possibly be provided by studies of similarly shaped cloud-form structures such as experimental convective plumes (Benech[14]), models of cumulus cloud formation (Squires and Turner[15]), and experimental jets (Hidy and Friedlander[16]; Morris[17]). Ejection

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11. Schlichting, H. (1968) Boundary Layer Theory, McGraw-Hill, New York, 744 pp.

12. Briggs, G.A. (1969) Plume Rise, AEC Critical Review Series USAEC, Report TID-25075, 81 pp.

13. Thorarinsson, S., and Vonnegut, B. (1964) Bull. Am. Meteorol. Soc. **45**: 440-444.

14. Benech, B. (1976) J. Appl. Meteor. **15**:127-137.

15. Squires, P., and Turner, J.S. (1962) Tellus **14**:422-434.

16. Hidy, G.M., and Friedlander, D.K. (1964) J. Am. Inst. Chem. Engr. **10**: 115-124.

17. Morris, D.G. (1968) Bull. Am. Meteor. Soc. **49**:1054-1058.

velocities observed during explosive volcanic eruptions (Chouet et al[18]) are typically much greater than updraft velocities produced by convective processes in the atmosphere. Eruption velocities correspond more closely to exit conditions at the mouth of a jet than to updraft velocities at the base of cumulus clouds or experimental thermal plumes. In addition, the temperature contrast between volcanic gases and the ambient atmosphere is more nearly approximated by some types of experimental jets (for example, Callaghan and Ruggeri[19]) than by updrafts associated with cumulus cloud formation.

The expansion of a turbulent jet in free flow (that is, unconfined by lateral boundaries) is determined by the rate at which the forward momentum of the jet is dissipated.[11, 20] Similarly the initial rise of an eruption cloud is principally determined by the upward momentum of dust and gas ejected from a volcanic vent. The internal structure of a turbulent jet may serve as a simple, first-order model of the internal structure of an explosive eruption cloud near the volcanic vent where the rise of dust and gas is controlled by the initial upward momentum of these materials. One method of extimating the atmospheric penetration of a turbulent volcanic jet is to compare the upward velocity of the jet with crosswind velocities above the volcanic vent at various altitudes. The initial upward momentum of the volcanic dust and gas can be considered to be effectively arrested at the altitude at which the vertical velocity of the jet, w, becomes comparable to the local crosswind velocity, u. Experimental studies of the actual behaviour of a turbulent jet in a crosswind have been reported by Keffer and Baines[21] and Patrick.[22]

The variation of vertical velocity w with range from a volcanic vent can be described by an expression for turbulent jet flow:[11]

$$w(x, z) = \frac{3}{8\pi} \frac{K}{\varepsilon_o z} \frac{1}{\left(1 + \frac{\eta^2}{4}\right)^2} \qquad (1)$$

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18. Chouet, B., Hamisevicz, N., and McGetchin, T.R. (1974) J. Geophys. Res. **79**:4961-4976.
19. Callaghan, E.E., and Ruggeri, R.S. (1948) Investigation of the Penetration of an Air Jet Directed Perpendicularly to an Air Stream, Report NACA-TN-1615, National Advisory Committee for Aeronautics.
20. Pai, S. (1954) Fluid Dynamics of Jets, Van Nostrand, New York, 221 pp.
21. Keffer, T.F., and Baines, W.D. (1963) J. Fluid Mech. **15**:481-497.
22. Patrick, M.A. (1967) Trans. Inst. Chem. Eng. (London) **45**:16-31.

where

$K = (1.59\ b\ w_o)^2$

$\varepsilon_o = 0.0256\ b\ w_o$

$\eta = \frac{1}{4\varepsilon_o} \sqrt{\frac{3K}{\pi}}\ \frac{x}{z}$

x = horizontal distance measured from jet centerline (meter)

z = vertical distance measured from the mouth of the vent (meter)

$w_o$ = average exit velocity at the vent (meter/sec)

b = half width of the jet, taken here as one-half the vent diameter (meter)

This equation is applicable to the region in which turbulent flow is fully developed, which normally occurs at a downstream range of approximately 10 vent diameters.[20] Along the centerline of the jet, (that is, directly above the volcanic vent) x = 0 and Eq. (1) reduces to

$$w(z) = \frac{3}{8\pi} \frac{K}{\varepsilon_o z} \qquad (2)$$

In order to apply this turbulent jet model to explosive eruption conditions, it is necessary to assume an average exit velocity for the erupted ash and gas. Quantitative descriptions of different types of explosive eruptions in terms of ejection velocities or mass flux rates are limited. Qualitatively, eruption intensity is considered to be related to the explosiveness of different types of eruptions.[10] Chouet et al[18] have observed gas exit velocities during individual explosive bursts of Strombolian-type eruptions that range from 110 m/sec to 20 m/sec. In the past, paroxysmal explosive eruptions have been accompanied by reports of repeated thunder and the firing of ships' guns at some distance from the actively erupting volcano, suggesting that gas exit velocity fluctuated around the speed of sound ~300 m/sec. Such reports occurred during the 1883 Krakatoa eruptions[23] and the 1902 eruption of Santa Maria Volcano in Guatemala.[24] Thus average exit velocities of approximately 300 m/sec may be tentatively associated with massive Plinian-scale eruptions.

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23. Symons, G. J. (1888) The Eruption of Krakatoa and Subsequent Phenomena, Report of the Krakatoa committee of the Royal Society, reprinted by Helio Associates, Inc., Tucson, Arizona, 1974.

24. Rose, W. I. (1972) Bulletin Volcanologique 36:29-45.

In this study the maximum height of an eruption cloud will be assumed to be related to the time-averaged eruption velocity of gas and fine ash at the mouth of a volcanic vent. In other words, fluctuations in eruption velocity are not considered to be important in determining the height of an eruption cloud. Average exit velocities ranging from 20 m/sec to 200 m/sec are assumed to represent a wide range of eruption intensity, from Strombolian to Vulcanian types of eruption. Much larger eruption velocities, on the order of 600 m/sec, have been inferred for the ballistic translation of large blocks of ejecta and for the formation of secondary craters commonly observed at distances of several kilometers from volcanic vents.[25] Such velocities are probably not representative of average exit conditions during an eruption but rather are associated with transient explosive pulses.

Figure 1 presents the variation of centerline jet velocity with altitude described by Eq. (2) for a circular vent with diameter $D = 100$ m over a range of eruption intensity (that is, different values of eruption velocity $w_o$). Also shown in Figure 1 is a series of vertical wind profiles representing averaged winter crosswind conditions in the Northern Hemisphere. These averaged wind profiles show that zonal westerly flow is stronger at mid-latitudes (Washington D.C. and Florida) than at subpolar latitudes (Greenland and the Aleutians).

At a particular altitude the centerline velocity along the jet represents the maximum upward velocity of any part of the eruption cloud. The maximum height to which volcanic dust and gas will rise as a result of their initial momentum can be approximately estimated as the altitude at which the jet centerline velocity equals the local crosswind velocity. Figure 1 indicates that for an eruption velocity of $w_o = 20$ m/sec (Strombolian-scale eruptions) the height of an eruption cloud may vary from $\Delta H \cong 1500 - 2000$ m at mid-latitudes to $\Delta H \cong 3500 - 4000$ m at subpolar latitudes; while for an eruption velocity of $w_o = 200$ m/sec (Vulcanian-scale eruptions) the height of an eruption cloud may vary from $\Delta H \cong 5000 - 6500$ m at mid-latitudes to $\Delta H \cong 11000 - 17000$ m at subpolar latitudes. These are maximum estimates of atmospheric penetration based upon the centerline velocity of the jet, not the average jet velocity at a particular altitude.

These cloud height estimates indicate that an order-of-magnitude increase in eruption velocity should result in a factor of 3 increase in the average height of an eruption cloud produced by a mid-latitude eruption, and a slightly larger factor of 4 increase in the average height of an eruption cloud produced by an eruption at subpolar latitudes. This method of estimating the height of an eruption cloud also indicates the important influence that crosswinds have on cloud height. For a particular value of eruption velocity the average height of an eruption cloud produced at subpolar latitudes is approximately twice the average height to which an eruption cloud would rise in the stronger westerly flow that occurs at mid-latitudes.

25. Fudali, R. F., and Melson, W. G. (1972) Bulletin Volcanologique 33:383-402.

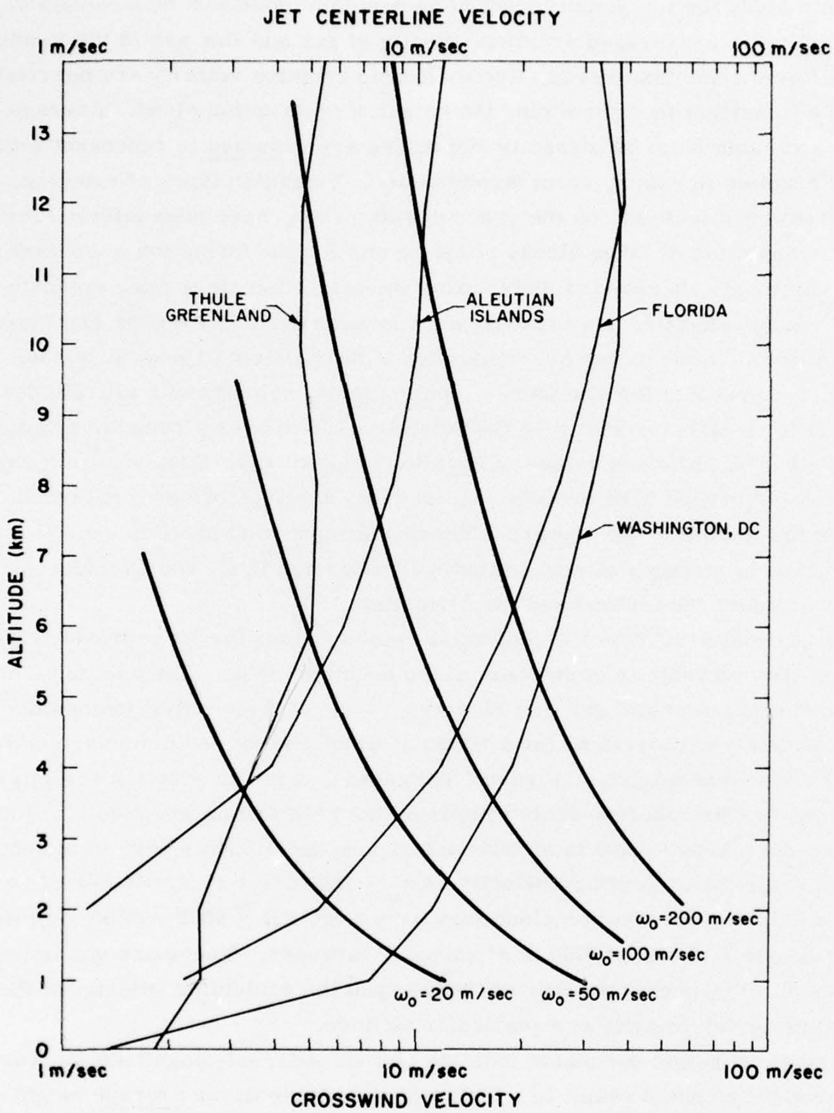


Figure 1. The Centerline Velocity of a Turbulent Volcanic Jet (heavy lines) Compared with Averaged Crosswind Velocities (light lines) at Various Altitudes. Jet centerline velocity is calculated by Eq. (2) for different values of $w_o$, the eruption velocity at a volcanic vent. Eruption velocities ranging from 20 m/sec to 200 m/sec are assumed to represent a wide variation in eruption intensity, from Strombolian-scale eruptions to Vulcanian-scale eruptions; a vent diameter D = 100 m has been assumed in all calculations. Wind profiles are averages for the winter season in the Northern Hemisphere. [Handbook of Geophys. and Space Environments (1965), Tables 4-12 through 4-18.]

Crosswinds can force an eruption cloud to bend over and become horizontal downrange of the volcanic vent. The actual trajectory of an eruption cloud in the presence of the prevailing crosswinds shown in Figure 1 can be roughly anticipated by the angle formed by the intersection of individual wind profiles and jet centerline velocity curves. The $w_o$ = 100 m/sec eruption velocity curve shown in Figure 1 converges at a small angle with the wind profile for Thule, Greenland at an altitude of ~12 km. A comparison of these two curves indicates that horizontal crosswind velocities are 70 percent as strong as the vertical centerline velocity of the volcanic jet over altitudes of 8 to 12 km. An eruption of this intensity into this crosswind environment would thus produce an eruption cloud that bends in a wide arc from the local vertical direction. In contrast the $w_o$ = 100 m/sec eruption velocity curve intersects the wind profile for Washington, D.C. at a much larger angle at an altitude of ~3500 m. In this case, the eruption cloud would bend through a much smaller arc in making the transition from predominantly vertical to predominantly horizontal motion.

Figure 1 indicates that the average atmospheric penetration of an eruption cloud produced by a turbulent volcanic jet should be greater for eruptions occurring at subpolar latitudes, where zonal westerly flow in the mid-troposphere is generally weaker than at middle latitudes. As mentioned previously, the average height of the tropopause is lowest near the poles (~9 km) so that direct introduction of volcanic dust and gas into the stratosphere may, on the average, be more easily accomplished by explosive eruptions at subpolar latitudes (for example, 1956 Bezymianny eruption in Kamchatka; 1912 Katmai eruption in Alaska). Similarly, zonal westerly flow in subtropical latitudes is weak in comparison with the mid-latitude westerlies. However, the average height of the tropopause is greatest in equatorial regions (~16 km). Thus massive Plinian-style eruptions are generally required to directly introduce volcanic dust and gas into the stratosphere at subtropical latitudes (for example, 1883 Krakatoa eruption in Indonesia; 1963 Mt. Agung eruption in Bali).

### 3.2 Rise of Industrial Plumes

Waste gases and fine particulate material released from industrial smokestacks form plumes that are sometimes clearly visible. The rate at which industrial effluents enter the atmosphere, though widely variable, is generally similar to some forms of fumarolic and weakly-explosive volcanic activity. The maximum gas discharge rates of commercial power plants are on the order of $10^3$ $m^3$/sec (Table 5.1, Ref 12) in comparison to a peak gas flux of $2 \times 10^3$ $m^3$/sec observed in the initial phases of individual explosive bursts from a volcanic vent at Stromboli by Chouet et al.[18] The rate at which thermal energy is released by such industrial

facilities is typically less than $5 \times 10^7$ cal/sec, whereas the gas phase transport of heat away from the surface of the permanent lava lake at the Nyirangongo Volcano has been estimated to be $10^8$ cal/sec by Delsemme[26] (see also Shimozuru[27]).

Eruption cloud heights of several hundred meters observed for Strombolian-style eruptions (see Appendix A) are comparable to the plume heights reported for a variety of industrial sources (see, for example Briggs[28]). In the case of major explosive eruptions (that is, Vulcanian-scale eruptions) exit conditions at volcanic vents differ significantly from those commonly found at the mouths of industrial smokestacks. Exit velocities and mass flux rates associated with major explosive eruptions greatly exceed the rate at which industrial effluents typically enter the atmosphere. (For example, Thorarinsson and Vonnegut[13] estimate that during the initial stages of the 1963 Surtesy eruption, thermal energy was emitted at a rate in excess of $10^{10}$ cal/sec.) As a result, eruption clouds produced by Vulcanian-style eruptions can easily dwarf most industrial plumes. The crosswind environment into which industrial effluents and volcanic dust and gas are emitted may also be quite different. Explosive eruptions commonly occur at the summits of large stratovolcanoes where crosswinds are likely to be considerably stronger than those typically encountered by industrial effluents.

A variety of theoretical and empirical expressions have been proposed to estimate the maximum rise heights of industrial plumes. Application of these industrially-based formulae to explosive eruption conditions involves an extrapolation of these various expressions beyond the range of exit conditions for which they were developed. As mentioned previously, there is an approximate correspondence between the heights of plumes produced by large industrial sources and the heights of eruption clouds produced by weakly-explosive Strombolian-style eruptions. At these scales vertical penetration of the atmosphere by industrial plumes and eruption clouds is roughly comparable. Extrapolation of industrially-based plume rise expressions beyond this scale to more explosive eruption conditions is employed here as a means of investigating the relationship between the heights of eruption clouds and eruption intensity. The rise of industrial plumes is predominantly controlled by the thermal buoyancy of industrial effluents. Estimates of the heights of eruption clouds based upon the behavior of industrial plumes can be compared with reported cloud heights and estimates of eruption cloud height based upon the turbulent jet model in order to assess the relative importance of thermal buoyancy in the rise of eruption clouds.

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26. Delsemme, A. (1960) Centre National de Volcan, Publ. **7**:699-707.
27. Shimozuru, D. (1968) Bulletin Volcanologique **32**:383-394.
28. Briggs, G.A. (1971) Nuclear Safety **12**:15-24.

The rise of industrial plumes emanating from smokestacks has been found to depend upon: (1) the velocity of the effluent at the mouth of the stack, (2) the temperature contrast between the effluent and the ambient atmosphere, (3) the cross-sectional area of the stack, (4) the average crosswind speed at the height at which the effluent is released, and (5) the thermal structure of the atmosphere (that is, the variation of environmental temperature with height). The following formulae employ various combinations of these parameters to estimate the maximum heights of industrial plumes (see also Figure 2). These formulae have been discussed in greater detail by Briggs.[29]

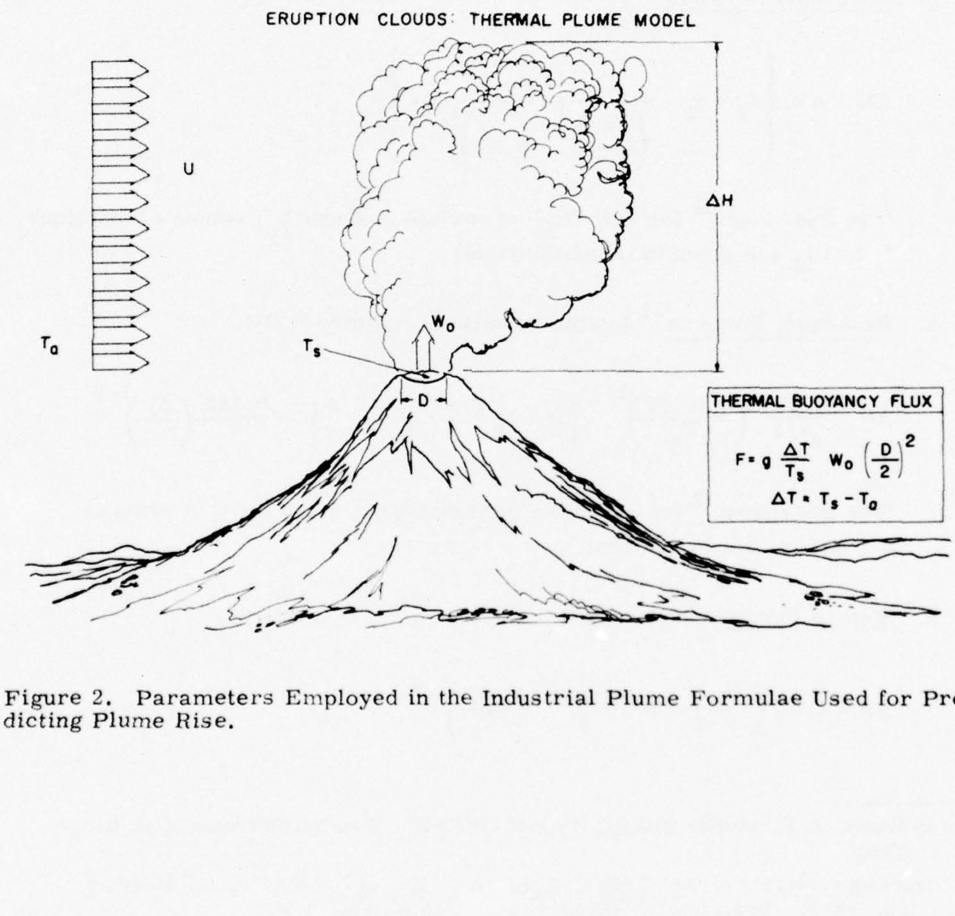


Figure 2. Parameters Employed in the Industrial Plume Formulae Used for Predicting Plume Rise.

29. Briggs, G.A. (1968) Momentum and buoyancy effects, in Meteorology and Atomic Energy, D. Slade (Ed), USAEC Report TID 24190.

1. Holland (Oak Ridge) Formula[30] (1953)

$$\Delta H = 1.5D\left(\frac{w_o}{u}\right) + 4.0 \times 10^{-5}\,\frac{Q_H}{u} \qquad (Q_H \text{ is heat flux in cal/sec})$$

2. Davidson-Bryant Formula[31] (1954)

$$\Delta H = D\left(\frac{w_o}{u}\right)^{1.4}\left(1 + \frac{\Delta T}{T_s}\right)$$

3. Bosanquet Formula[32] [Stable conditions, windy (1957)]

$$\Delta H = A\,u\left\{ f_1 + f_2 - \frac{0.615\,X_o^{1/2}}{\left(\left(\frac{w_o}{u}\right)^2 + 0.57\right)^{1/2}} \right\}$$

(See Bosanquet[32] for definition of variables A and $X_o$; values of functions $f_1$ and $f_2$ are given in tabular format)

4. Bosanquet Formula[32] [Stable conditions, calm (1957)]

$$\Delta H = \frac{0.666}{\alpha^{1/2}}\left(\frac{g\,Q\,\Delta T}{T_a}\right)^{1/4}\left\{(t + t_o)^{3/4} - \frac{1}{2}\,t_o^{3/4}\right\} - \frac{0.283}{\alpha}\left(\frac{Q}{w_o}\right)^{1/2}$$

(See Bosanquet[32] for definition of variables $\alpha$, t and $t_o$; Q is effluent discharge rate in $m^3$/sec)

5. Stümke Formula[33] (1963)

$$\Delta H = 1.5D\left(\frac{w_o}{u}\right) + 65.0\,\frac{D^{3/2}}{u}\left(\frac{\Delta T}{T_s}\right)^{1/4}$$

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30. Holland, J. Z. (1953) USAEC Report ORO-99, Weather Bureau, Oak Ridge Tenn.

31. Davidson, W. F. (1954) Trans. Conf. Ind. Wastes, 14th Annual Meeting, pp. 38-55, Industrial Hygiene Foundation of America.

32. Bosanquet, C. H. (1957) J. Inst. Fuel 30:322-328.

33. Stümke, H. (1963) U.S. Atomic Energy Commission Report ORNL-TR-977, Oak Ridge National Laboratory.

6. Briggs Formula[12] [Stable conditions, windy (1969)]

$$\Delta H = 4.0 \left(\frac{F}{uS}\right)^{1/3}$$

7. Briggs Formula[12] [Stable conditions, calm (1969)]

$$\Delta H = 5.0 \frac{F^{1/4}}{S^{3/8}}$$

*where*

$\Delta H$ = plume rise height above vent (m)

$D$ = vent diameter (m)

$w_o$ = vertical exit velocity at vent mouth (m/sec)

$u$ = average crosswind speed (m/sec)

$\Delta T$ = difference in absolute temperature between ambient air and effluent gas ($^oK$)

$T_s$ = absolute temperature of effluent stack gas ($^oK$)

$T_a$ = absolute temperature of ambient atmosphere ($^oK$)

$F$ = buoyancy flux $F = g\left(\frac{\Delta T}{T_s}\right) w_o \left(\frac{D}{2}\right)^2$, ($m^4/sec^3$)

$S$ = atmospheric stability parameter $= \frac{g}{T_a} \frac{\partial \Theta}{\partial Z}$, ($1/sec^2$)

$$\frac{\partial \Theta}{\partial Z} = \Gamma + \frac{9.8^oK}{1000\ m}$$

$\Gamma$ = environmental lapse rate $= \partial T/\partial Z$ ($^oK/m$)

$g$ = gravitational acceleration (9.8 $m/sec^2$)

These various formulae are based upon both laboratory experiments and observations of the actual behavior of industrial plumes. Each formula produces reasonably accurate estimates of plume rise when applied to certain types of meteorological conditions and a specific range of effluent source strength. No single technique can accurately predict plume height in all cases. For example, the Holland (Oak Ridge) formula is based upon wind tunnel experiments and empirical observations at relatively small power plants operating in the 1950's. When applied to explosive eruption conditions the Holland (Oak Ridge) formula predicts unreasonably large cloud heights. (For values of $Q_H$ greater than $10^{10}$ cal/sec the Holland formula

predicts eruption cloud heights on the order of 50 km.) The remaining equations presented above produce realistic estimates of eruption cloud height when applied to explosive eruption conditions assumed in this study. By considering several methods of estimating plume height it is possible to make a meaningful estimate of eruption cloud height based upon the behavior of industrial plumes.

Table 1 presents the calculated heights of eruption clouds above an actively-erupting volcanic vent for the case of a volcanic gas ($T_s = 373^oK$) entering the ambient atmosphere ($T_a = 273^oK$) via a circular vent of diameter D = 100 m. It has been assumed here that the erupted gases expand rapidly and cool from the temperature at which the magma is erupted to 100°C such that they effectively exit the crater at the latter temperature. Small changes in $T_a$ or $T_s$ will not greatly affect cloud height estimates in Table 1. These estimates have been rounded to the nearest 100 m in recognition of the large extrapolations involved in applying the industrial plume formulae to explosive eruption conditions. Actual observations of the heights of eruption clouds are commonly estimated to the nearest kilometer (see Appendix A). A variety of combinations of eruption velocity and crosswind velocity has been chosen to represent an increase in eruption intensity from the left to right of Table 1. Explosive eruptions typically occur at the summits of large stratovolcanoes where average crosswind velocities are on the order of 10 to 30 m/sec.

Estimated heights of eruption clouds vary from 900 - 10,000 m for Strombolian-scale eruptions ($w_o$ = 20 m/sec) to 3200 - 8500 m for Vulcanian-scale eruptions ($w_o$ = 200 m/sec) in Table 1. The Stümke formula appears to predict unreasonably large cloud heights ($\Delta H \sim 10$ km) for relatively small, Strombolian-scale eruptions ($w_o$ = 20 m/sec); and the Briggs and Bosanquet formulae for calm conditions may be considered to be inappropriate for the upper levels of the atmosphere where crosswind velocities are usually greater than 5 m/sec. A more selective range of estimates can be based upon the Davidson-Bryant, Bosanquet (stable conditions, windy) and Briggs (stable conditions, windy) formulae. Disregarding the Stümke formula, estimates of cloud height in the presence of a crosswind vary from 900 - 4000 m for Strombolian-scale eruptions ($w_o$ = 20 m/sec) to 3200 - 8400 m for Vulcanian-scale eruptions ($w_o$ = 200 m/sec). Based upon these estimates, it would appear that an order of magnitude variation in eruption velocity will not necessarily result in a major change in eruption cloud height. A comparison of the median value of cloud height selected from the range of estimates predicted for these two cases of eruption conditions suggests that an order of magnitude increase in eruption intensity should result in a factor of 2.5 increase in median cloud height.

A comparison of the Davidson-Bryant, Bosanquet (stable conditions, windy), and Briggs (stable conditions, windy) formulae for an eruption velocity of $w_o$ = 200 m/sec under different crosswind conditions indicates that cloud height varies from

Table 1. Eruption Cloud Height Estimates* [Based upon the industrial plume models discussed in the text for various combinations of eruption velocity ($w_o$) and crosswind velocity (u).]

| CROSSWIND CONDITIONS | | | | | | | |
|---|---|---|---|---|---|---|---|
| Eruption velocity | | $w_o$ = 20 m/sec | $w_o$ = 50 m/sec | $w_o$ = 100 m/sec | $w_o$ = 200 m/sec | $w_o$ = 200 m/sec | $w_o$ = 300 m/sec |
| Crosswind velocity | | u = 5 m/sec | u = 10 m/sec | u = 10 m/sec | u = 10 m/sec | u = 20 m/sec | u = 10 m/sec |
| Davidson-Bryant | | 900 m | 1200 m | 3200 m | 8400 m | 3200 m | 14800 m |
| Bosanquet | $\Gamma$ = -7 | 2400 | 2700 | 3200 | 3700 | 3500 | 3800 |
| (stable) | $\Gamma$ = -9 | 4000 | 4400 | 5400 | 6300 | 5200 | 6800 |
| Stümke | | 10000 | 5400 | 6200 | 7700 | 3800 | 9200 |
| Briggs | $\Gamma$ = -7 | 2600 | 2800 | 3500 | 4400 | 3500 | 5000 |
| (stable) | $\Gamma$ = -9 | 3900 | 4200 | 5300 | 6600 | 5300 | 7600 |
| CALM CONDITIONS | | | | | | | |
| Eruption velocity | | $w_o$ = 20 m/sec | $w_o$ = 50 m/sec | $w_o$ = 100 m/sec | $w_o$ = 200 m/sec | $w_o$ = 200 m/sec | $w_o$ = 300 m/sec |
| Bosanquet | $\Gamma$ = -7 | 2500 m | 3100 m | 3500 m | 3700 m | 3700 m | 3700 m |
| (stable) | $\Gamma$ = -9 | 4200 | 5200 | 6000 | 6700 | 6700 | 7100 |
| Briggs | $\Gamma$ = -7 | 3000 | 3800 | 4500 | 5300 | 5300 | 5900 |
| (stable) | $\Gamma$ = -9 | 4800 | 6000 | 7200 | 8500 | 8500 | 9500 |

*Note: Eruption velocity increases *from* left to right. Calculated cloud heights have been rounded to the nearest 100 m. Gamma ($\Gamma$) is the environmental lapse rate, $\partial T/\partial Z$, in units of $^{o}K/km$; larger values of $\Gamma$ represent increasingly stable atmospheric conditions.

$\Delta H = 3700 - 8400$ m for crosswind $u = 10$ m/sec to $\Delta H = 3200 - 5300$ m for crosswind $u = 20$ m/sec. Thus variations in crosswind conditions may potentially be as significant as variations in eruption velocity in determining the height of an eruption cloud. In addition, Table 1 demonstrates that variations in the thermal structure of the atmosphere ($\Gamma$) can also have an important influence on eruption cloud height.

## 4. DISCUSSION

The physical mechanisms responsible for the rise of an eruption cloud are the upward momentum and thermal buoyancy of the volcanic dust and gas. Each of the analogous phenomena employed here to model the rise of an eruption cloud is principally dependent on one of these two physical mechanisms. By comparing cloud height estimates produced by these two very different models, it may be possible to infer which of the two mechanisms plays a more important role in the rise of volcanic eruption clouds.

Estimates of eruption cloud height based upon the turbulent jet model are a measure of atmospheric penetration produced by the upward momentum of volcanic dust and gas. Cloud height estimates based upon the industrial plume models are a measure of atmospheric penetration produced by the thermal buoyancy of volcanic gas. For eruption velocities varying from 20 to 200 m/sec, cloud heights estimated by the turbulent jet model range from 1500 - 6500 m (mid-latitude eruption in Figure 1), whereas cloud heights estimated by the industrial plume models range from 900 - 8400 m (preferred models neglecting Strümke Formula and calm conditions in Table 1). These estimates of cloud height are considered to be roughly comparable in view of the assumptions and extrapolations involved in applying the models to explosive eruption conditions. The fact that generally similar estimates of cloud height are produced by two very different models suggests that both momentum and thermal buoyancy play an important role throughout the main portion of an eruption cloud's trajectory. For these eruption conditions ($20$ m/sec $\leq w_o \leq 200$ m/sec) neither momentum nor thermal buoyancy appears to dominate the process of cloud rise to altitudes of $\sim 10$ km above an actively-erupting volcanic vent.

Both the turbulent jet and the industrial plume models indicate that for constant crosswind conditions the height of an eruption cloud should increase as eruption intensity (that is, average eruption velocity $w_o$) increases. An order of magnitude variation in eruption velocity from $w_o = 20$ m/sec to $w_o = 200$ m/sec results in a factor of 3 to 4 increase in average cloud height predicted by the turbulent volcanic jet model and a factor of 2.5 increase in median cloud height predicted by the select group of industrial plume models discussed earlier. However, both

models also demonstrate that changes in crosswind velocity by factors of 2 to 5 can result in variations in cloud height of similar magnitude. The models indicate that a volcanic cloud produced by a large-scale explosive eruption in a strong crosswind environment could rise to smaller heights than a volcanic cloud produced by a moderate-scale explosive eruption in a weak crosswind environment. Therefore, the apparent heights of eruption clouds do not necessarily reflect relative eruption intensity. Reported heights of eruption clouds without reference to local crosswind conditions at the time of eruption cannot be directly compared to gauge the relative explosiveness of different volcanic eruptions.

The range of eruption cloud heights predicted by the turbulent jet and the industrial plume models for average eruption velocities varying from 20 to 200 m/sec is generally less than 10 km. This agrees quite well with the range of observed eruption cloud heights reported to the Center for Short Lived Phenomenon over the period 1970-1974 (see Appendix A). Such eruptions may introduce large quantities of particulate dust and volcanic gas regionally into the troposphere but do not directly penetrate the tropopause. Volcanic dust and gas produced by eruptions of this size may nevertheless enter the stratosphere via stratosphere-troposphere exchange processes.[34]

Maximum observed eruption cloud heights reported over the period 1970-1974 ranged up to 12 km for the February 1973 eruption of Fuego Volcano, Guatemala and 15 km for the summer 1970 eruption of Hekla, Iceland (see Appendix A). Both of these eruptions occurred outside the $30^{\circ}$-$60^{\circ}$ latitude band where zonal westerly flow is best developed. The Hekla eruption should have injected material directly into the stratosphere. Widespread stratospheric effects produced by the Fuego eruption have been reported by several investigators.[35, 36, 37, 38]

Even greater eruption cloud heights ranging up to 20 km to 40 km have been inferred or reported for catastrophic Plinian-scale explosive eruptions in the past (Lamb, 1970; Cronin, 1971). Eruption velocities greater than 200 m/sec are required by both the industrial plume models and the turbulent jet model in order to predict eruption cloud heights greater than 20 km under reasonable crosswind conditions ($u \geq \sim 5$ m/sec). Presumably such average eruption velocities must be maintained for some period of time (on the order of several hours ?) to permit the eruption cloud to attain its maximum height. First hand measurements of eruption velocities and crosswind conditions during such catastrophic explosive eruptions

34. Reiter, E. R. (1975) Revs. Geophys. and Space Sci. 13:459-474.
35. Fegley, R., and Ellis, H. T. (1975) Geophys. Res. Lett. 2:139-141.
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are understandably rare. Therefore it is difficult to determine if estimates of average eruption velocity on the order of 300 m/sec - 500 m/sec, inferred by comparing actual eruption cloud heights with model predictions, are reasonable for these massive Plinian-scale eruptions.

The rise of such gigantic eruption clouds may also be influenced by:

(1) Atmospheric Thermal Structure. Eruption clouds rising above 10 km may be dominantly controlled by runaway convection in an atmosphere where the environmental lapse rate was slightly greater than the adiabatic lapse rate (that is, unstable conditions). In such an environment, the introduction of volcanic dust and gas could trigger large-scale convection. Observations of the behavior of industrial plumes in neutral or unstable atmospheric conditions are limited.[12] It is difficult to estimate exactly how much farther an eruption cloud might rise in such adiabatic or superadiabatic conditions.

Thermal inversion layers can restrict convective motion in the atmosphere. Inversions are produced by heating of the earth's surface and are typically confined to the lower 2 km of the atmosphere. At such altitudes the upward momentum of volcanic dust and gas ejected by an explosive eruption should generally be sufficient to carry an eruption cloud through an inversion layer (see Figure 1). Therefore, the presence of inversion layers in the lower atmosphere should not usually influence the rise of explosive eruption clouds.

(2) Anomalously Calm Upper Level Winds. As discussed above, both the industrial plume models and the turbulent jet model are sensitive to variations in crosswind strength at the time of an eruption. Unusually calm conditions aloft ($u < \sim 5$ m/sec) will permit eruption clouds produced by eruption velocities less than 200 m/sec to attain altitudes greater than $\sim 15$ km according to both types of models presented here. Such calm conditions may occur as the result of a fortuitous configuration of tropospheric weather systems at the time of an eruption.

(3) Sustained Eruption Conditions. An individual eruption can pass through a series of phases alternating between Strombolian and Vulcanian styles of eruption, occasionally culminating in a massive Plinian-scale eruption ($w_o \gtrsim 300$ m/sec ?). During periods of sustained eruption, the eruption cloud produced by the initial explosive phases of activity may establish a natural vertical conduit in the atmosphere. It is possible that gas and dust released during a short-lived increase in eruption intensity may rise through the pre-existing cloud column to produce a significant increase in the maximum height of an eruption cloud. The increase in cloud height produced by the transient, more explosive phase may persist through later eruptive phases. Thus, the observed maximum height of an eruption cloud may depend to an important degree on the chronological sequence of eruptive phases produced during a single period of eruption, and not necessarily on the time-averaged eruption conditions.

## 5. CONCLUSIONS

1. Turbulent jet and industrial plume models of volcanic eruption clouds predict comparable rise heights over a range of eruption conditions for explosive eruptions occurring in the mid-latitudes. The atmospheric penetration of a turbulent jet is principally due to the momentum of the ejected materials; the atmospheric penetration of an industrial plume is principally due to the thermal buoyancy of industrial effluents. The fact that comparable cloud height estimates are produced by these two very different models suggests that both momentum and thermal buoyancy play an important role throughout the main portion of an eruption cloud's trajectory.

2. Estimated heights of eruption clouds based upon turbulent jet and industrial plume models are sensitive to assumed crosswind conditions. Both types of models indicate that variations in eruption intensity and crosswind velocity can result in changes in cloud height that are of similar magnitude. Reports of observed eruption cloud heights rarely include information about local meteorological conditions, frequently due to the remote location of many major explosive eruptions. The height of an eruption cloud has in the past been regarded as a relative index of eruption intensity, especially in cases where ground observations of the actual eruption are difficult or sparse. This study indicates that the heights of eruption clouds do not necessarily reflect relative eruption intensity (measured here as average exit velocity at the volcanic vent). Observations of eruption cloud height may yet provide a relative index of eruption intensity if they can be reported with reference to local crosswind conditions at the time of the eruption. Knowledge of the thermal structure of the atmosphere would also be useful in relating observed heights of eruption clouds to surface eruption conditions.

The turbulent jet and industrial plume models also indicate that either the momentum or the thermal buoyancy of volcanic dust and gas can carry an eruption cloud to heights on the order of 5 to 10 km. Clouds rising to similar heights may thus be produced by large areal sources of hot volcanic gas ejected at small eruption velocities and small, violently-explosive sources of volcanic gas and dust ejected at large eruption velocities. Therefore, even if local crosswind conditions are noted, the apparent heights of eruption clouds may not always provide a useful index to relative eruption intensity.

3. The turbulent jet/industrial plume models can be used to infer minimum eruption velocities for catastrophic explosive eruptions which have in the past produced clouds rising to heights of 20 km or more (for example, the 1883 Krakatoa eruption). For reasonable crosswind conditions ($u > \sim 5$ m/sec) both types of models require eruption velocities greater than 200 m/sec in order to predict cloud heights greater than 20 km. The formation of such gigantic eruption clouds

may also depend upon: (a) environmental lapse rates in the upper troposphere which favor large-scale convection, (b) anomalously calm crosswind conditions in the upper troposphere, or (3) transient increases in eruption intensity during periods of sustained explosive activity.

4. Future investigations of explosive volcanic eruptions should include reports of local meteorological conditions during the eruption where possible. Daily observations of ground-based terms should include measurements of air temperature, surface pressure, wind speed, and the surface concentration of volcanic dust and gas. Daily aerial observations should include measurements of the height and external structure of the cloud, radial temperature and pressure variations about the cloud, local crosswind profiles and lapse rates, and volcanic gas and dust concentrations in the vicinity of the cloud. Simultaneous observations of this nature could be used to model the structure of such clouds in order to determine the manner in which they introduce volcanic dust and gas into the earth's atmosphere. Such knowledge would provide an important point of reference in assessing the relative impact of anthropogenic sources of effluents which are episodically injected directly into the upper levels of the atmosphere.

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# Appendix A

## Observed Eruption Cloud Heights

Reported heights of volcanic eruption clouds come from a variety of sources. The following listing of observed eruption cloud heights has been compiled from a series of Annual Reports of the Center for Short-Lived Phenomena covering the period 1970-1974 and supplemented using other sources where noted. (In cases where a variety of cloud heights has been reported for a single period of eruption at an individual volcano the maximum reported height has been included in this listing.) This listing is not necessarily comprehensive; rather it indicates the range of eruption cloud heights that are actually observed in nature. It is frequenctly difficult to determine if reported cloud heights have been measured with respect to sea level, the elevation of the erupting vent, or the elevation of the observer. Reports of cloud height less than 1000 m usually are made by observers near the erupting volcanic vent and typically refer to height above the vent. Reports of cloud height greater than 10,000 m are usually made by ground-based or airborne observers at some distances from the erupting vent. Aerial observations typically refer to height above sea level, whereas ground-based observations commonly refer to height above the observer. Tabulations of observed and inferred eruption cloud heights reported before 1970 have been presented by Lamb and Cronin.[8, 9]



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Table A1. Observed Eruption Cloud Heights

| Approximate Date of the Eruption | Location | Reported Height of the Eruption Cloud (meters) |
|---|---|---|
| 25 Oct 1969 | Bezymianny Volcano, Kamchatka Peninsula, USSR | 2000 |
| 27 Nov 1969 | Mt. Lokon, Indonesia | 500 |
| 27 Dec 1969 | Cerro Negro, Nicaragua | 700 |
| 30 Mar 1970 | San Miguel Volcano, El Salvador | 400 (above the crater floor) |
| 21 Apr 1970 | Mt. Aso, Japan | 150 (above the crater floor) |
| 5 May-5 Jul 1970 | Hekla, Iceland | 15, 000 |
| 11 May 1970 | Karimsky Volcano, Kamchatka Peninsula, USSR | 8000 |
| 28 May 1970 | Suwanosezima Volcano, Japan | 2000-3000 |
| 25 Aug 1970 | Telica Volcano, Nicaragua | 500 |
| Sep-Dec 1970 | Akita-Komagatake Volcano, Japan | 400 |
| 20 Sep 1970 | Beerenberg Volcano, Jan Mayen Island, Greenland Sea | 5000-6000 |
| 16 Nov 1970 | Sakurazima Volcano, Japan | 2100 |
| 11 Jan 1971 | Sakurazima Volcano, Japan | 3000 |
| 3 Feb 1971 | Cerro Negro Volcano, Nicaragua | 6100 |
| 16 Feb 1971 | Sakurazima Volcano, Japan | 1800 |
| 10 Apr 1971 | Sakurazima Volcano, Japan | 2000 |
| Sep 1970-June 1971 | Akita-Komagatake Volcano, Japan | 600 (above the crater) |
| 19 Jul 1971 | White Island Volcano, North Island, New Zealand | 2000 |
| 18 Aug 1971 | Mt. Hudson, Chile | 7000 above sea level ~ 5000 meters above summit) |
| 18-22 Jun 1972 | Alaid Volcano, Kuril Island, USSR | 4000-8000 |
| 21 Jun 1972 | Poas Volcano, Costa Rica | 3000 |
| 13-15 Sep 1972 | Sakurazima Volcano, Japan | 3000-4000 |
| 6 Oct 1972 | Mt. Merapi, Java, Indonesia | 3000 |
| Dec 1972 | Anak Krakatoa, Indonesia | 1600 |
| 23 Jan-3 Jul 73 | Helgafell Volcano, Iceland | not reported |

Table A1. Observed Eruption Cloud Heights (Continued)

| Approximate Date of the Eruption | Location | Reported Height of the Eruption Cloud (meters) |
|---|---|---|
| 22 Feb 1973 | Fuego Volcano, Guatemala | 12,000 |
| 18 Apr 1973 | Asama Volcano, Japan | 4500 |
| 1 May 1973 | Long Island, New Guinea | 150 |
| Jun 1973 | Sakurazima Volcano, Japan | 5000 |
| 14 Jul 1973 | Tiatia Volcano, Kurile Islands, USSR | 5500 |
| 14 Jul 1973 | Curacoa Reef Submarine Eruption, Tonga Islands | 5000 |
| 16 Sep 1973 | Santiaguito Volcano, Guatemala | 8000 |
| 1974 | Sakurazima Volcano, Japan | 2000-3300 |
| 23-28 Feb 1974 | Reventador Volcano, Ecuador | 1000 |
| 14 Oct 1974 | Fuego Volcano, Guatemala | 7000 (Bonis, 1976) |
| 15 Oct 1974 | Klyuchevskoy Volcano, Kamchatka Peninsula, USSR | 2000-3000 |